A High-Resolution Simulation of Asymmetries in Severe Southern Hemisphere Tropical Cyclone Larry (2006)

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ABSTRACT

Advances in observations, theory, and modeling have revealed that inner-core asymmetries are a common feature of tropical cyclones (TCs). In this study, the inner-core asymmetries of a severe Southern Hemisphere tropical cyclone, TC Larry (2006), are investigated using the fifth-generation Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model (MM5) and the Kepert–Wang boundary layer model. The MM5-simulated TC exhibited significant asymmetries in the inner-core region, including rainfall distribution, surface convergence, and low-level vertical motion. The near-core environment was characterized by very low environmental vertical shear and consequently the TC vortex had almost no vertical tilt. It was found that, prior to landfall, the rainfall asymmetry was very pronounced with precipitation maxima consistently to the right of the westward direction of motion. Persistent maxima in low-level convergence and vertical motion formed ahead of the translating TC, resulting in deep convection and associated hydrometeor maxima at about 500 hPa. The asymmetry in frictional convergence was mainly due to the storm motion at the eyewall, but was dominated by the proximity to land at larger radii. The displacement of about $30^\circ$–$120^\circ$ of azimuth between the surface and midlevel hydrometeor maxima is explained by the rapid cyclonic advection of hydrometeors by the tangential winds in the TC core. These results for TC Larry support earlier studies that show that frictional convergence in the boundary layer can play a significant role in determining the asymmetrical structures, particularly when the environmental vertical shear is weak or absent.

1. Introduction

Tropical cyclones (TCs) are among the most destructive of all weather-related natural hazards. The inner-core region of a TC that surrounds the relatively calm eye consists of an annulus of very intense winds [i.e., the radius of maximum winds (RMW)], which produces the most severe wind-related structural damage upon landfall. Very heavy rainfall often occurs in the inner core as a result of strong updrafts within the eyewall. Because of these and other related hazards, it is important to understand the processes that control the inner-core structure of TCs.

The structure of a mature TC can be highly asymmetrical, depending on both the complex interaction between the environment and the TC, and the internal dynamics of the TC itself. Several mechanisms have been proposed to explain the asymmetries observed and simulated in TCs. A number of studies have emphasized the importance of the environmental vertical wind shear in determining the spatial distribution of deep vertical
motion and resultant precipitation in TCs. Black et al. (2002) documented the observed structures of Hurricanes Jimena (1991) and Olivia (1994) with respect to changes in the environmental vertical shear. They showed that for relatively weak shear of less than 5 m s$^{-1}$, the convection around the eyewall in Hurricane Olivia was symmetrically distributed. For vertical shear values above 8 m s$^{-1}$, the distribution of convection became asymmetrical, characterized by a dominant wavenumber-1 component, with enhanced reflectivities in the semicircle to the left of the shear vector. The asymmetry of Hurricane Olivia was investigated also by Reasor et al. (2000). The vertical alignment of the vortex and distribution of convection was found to be sensitive to a sudden increase in the environmental vertical shear (from 3–5 to 15 m s$^{-1}$ in 2.5 h), with the inner core developing a downshear tilt after being initially vertically aligned. A double maximum in inner-core convection developed in the downshear direction with the largest vertical motion in the downshear-left quadrant. Frank and Ritchie (1999, 2001) showed that for TC-like vortices on an f plane, vertical shear played a fundamental role in the location of convection within the core region, with maximum vertical motion occurring downshear and to the left of the vertical shear vector. The simulated vortices were found to tilt in the direction of the vertical wind shear and this tilt was associated with asymmetries of vertical motion. The resulting rainwater was advected cyclonically to the left of the shear vector such that the accumulated rainfall was greater to the left of the direction of the storm motion. Braun (2002) and Braun et al. (2006) performed high-resolution simulations of Hurricane Bob (1991) and Hurricane Bonnie (1998), respectively, and found marked wavenumber-1 asymmetries in the inner-core region with maximum vertical motions and boundary layer inflow located in the direction of vortex tilt, associated with the deep layer vertical wind shear. Rogers et al. (2003) also investigated the role of vertical shear in the precipitation structure of Hurricane Bonnie; they found that the magnitude of the vortex tilt correlated well with the magnitude of vertical shear and that the accumulated rainfall pattern tended to be asymmetric and to the left of the track when the shear was relatively weak and aligned with the storm motion. Braun et al. (2006) found that strong updrafts within the TC eyewall formed and moved in conjunction with lower-tropospheric eyewall mesovortices, and accounted for a large portion of the total upward mass flux. Similarly, Braun and Wu (2007) in an investigation of the inner-core structure of Hurricane Erin (2001) concluded that the organization of eyewall vertical motion was influenced strongly by the strength of the vertical wind shear, with relatively weak shear (<5 m s$^{-1}$) allowing mesovortices (and associated convective updrafts) to form on all sides of the eyewall. As the shear increased, the convective-scale updrafts tended to form predominantly on the downshear side of the eyewall and then decay as they moved upshear.

While the environmental vertical wind shear has been shown to play a fundamental part in explaining the dynamics of asymmetries in TCs, several studies have demonstrated that asymmetries can develop also in the absence of vertical shear because of the asymmetric friction associated with a translating vortex. Shapiro (1983) used a depth-averaged boundary layer model to show that enhanced low-level convergence occurred ahead of the TC, relative to its motion, due to this motion-induced frictional asymmetry. Kepert (2001) and Kepert and Wang (2001) found similar results using full three-dimensional boundary layer models with analytic and numerical solutions, respectively. Bender (1997), using a full tropical cyclone model, found that for a TC embedded in a constant basic easterly flow (5 m s$^{-1}$) with no shear, maximum boundary layer convergence occurred at the front side of the storm, where the relative inflow was also maximized.

The relative contributions of both storm motion and environmental shear in determining convective asymmetries in TCs has been investigated by Corbosiero and Molinari (2003) and Frank and Ritchie (1999). Based on observed lightning data, Corbosiero and Molinari (2003) demonstrated that for slow (<3 m s$^{-1}$), moderate (3–6 m s$^{-1}$), and fast-moving (>6 m s$^{-1}$) storms, the highest flash counts in the inner core occurred in the right-front quadrant relative to the direction of motion. When the direction of the vertical shear vector was opposite to the direction of motion, the highest flash counts occurred mostly in the direction of shear and therefore it was argued that the shear effect tends to dominate the motion effect in determining the convective asymmetries.

The land–sea contrast may also impact the storm symmetric structure, although this has not been as extensively studied as the effects of motion and shear. Boundary layer aspects of this problem were first considered by Kepert (2002a,b), who noted an analogy with the motion-induced asymmetry. This idea was extended in the observation- and model-based studies of Hurricane Mitch (1998; Kepert 2006b) and severe Tropical Cyclone Ingrid (May et al. 2008), which further noted a strong convective asymmetry although it was not possible to separate the effects of shear and proximity to land in those papers. Wong and Chan (2004) presented a more systematic study and confirmed that nearby land can induce a marked asymmetry in the low-level flow.
The aim of this study is to document the inner-core asymmetries of a severe Southern Hemisphere TC (TC Larry in March 2006) in relation to the two primary mechanisms discussed above, using data previously obtained from a high-resolution numerical simulation (Ramsay and Leslie 2008). TC Larry formed in the southwest Pacific Ocean over the Coral Sea region east of Australia at 0600 UTC 17 March 2006. It intensified fairly rapidly as it moved westward, eventually making landfall as an Australian severe category 4 TC (wind gusts between 63 and 78 m s$^{-1}$) on the northeast Australian coast near the town of Innisfail. Unlike many U.S. hurricanes, the detailed structure of TC Larry’s inner core remains largely unknown because of the absence of aircraft measurements of TCs in the Australian region. However, analyses of passive microwave imagery at 85 GHz during the 12 h preceding landfall (0900–2100 UTC 19 March 2006) revealed a pronounced convective asymmetry with enhanced convection both on the western and northern sides of a well-defined eyewall (see, e.g., the microwave imagery of TC Larry at the Naval Research Laboratory tropical cyclone Web site http://www.nrlmry.navy.mil/tc_pages/tc_home.html). This convective asymmetry is clearly evident in one of the passes from the Tropical Rainfall Measuring Mission (TRMM) satellite at 1001 UTC (Fig. 1)—about 10.5 h before landfall. The satellite-derived polarization corrected temperature (PCT) can be used to detect regions of deep convection within the TC core, with colder brightness temperatures indicating more intense convection due to the scattering of upwelling radiation by ice particles above the freezing level. Finally, TC Larry passed within 200 km of the Willis Island radar site (−16.29°S, 149.97°E), about 450 km east of Cairns, Australia, from 0100 to 1000 UTC 19 March. This radar imagery revealed persistent maxima in low-level reflectivity (height ~1500 m) on the northern side of well-defined eyewall (not shown).

This article is organized as follows: section 2 describes in detail the model configurations and numerical experiments. The results are presented in section 3, and section 4 provides a discussion and summary of our findings.
2. Description of numerical experiments and techniques

The primary simulations were carried out with version 3.7 of the fifth-generation Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model (MM5). A detailed description of MM5 is given in Grell et al. (1994).

A quadruply nested, two-way interactive grid was used to simulate TC Larry, with the finest grid spacing of 1 km used in the innermost nest as shown in Fig. 2. The four domains have dimensions and grid spacings of $93 \times 100$, 27 km (D1); $220 \times 210$, 9 km (D2); $445 \times 286$, 3 km (D3); and $385 \times 268$, 1 km (D4) (Fig. 2). In total, 46 vertical half-sigma levels were used with higher resolution in the boundary layer and upper troposphere to more accurately resolve the inflow and outflow layers of the TC. The model top was set at 50 hPa. The simulation was initialized at 1200 UTC 17 March and run for 72 h until the TC decayed roughly to the Australian category 1 strength (i.e., gusts to $35 \text{ m s}^{-1}$) about 12 h after landfall. The initial and lateral boundary conditions were provided by the National Centers for Environmental Prediction (NCEP) Final Analysis (FNL) dataset that has $1^\circ \times 1^\circ$ horizontal grid spacing and 26 mandatory pressure levels from 1000 to 10 hPa. A bogus vortex was inserted at $t = 0$ following the TC bogusing scheme of Low-Nam and Davis (2001). The model failed to generate a TC when no bogus vortex was used because of the relatively coarse resolution of the FNL dataset. The vertical shear of the environmental wind was computed as the vector difference between winds at 200 and 850 hPa, averaged over a $5^\circ \times 5^\circ$ box centered on the eye of the TC using data from D3. Similarly, we define a local vertical shear vector using the same pressure levels (850–200 hPa) and center but averaged over a much smaller, $0.4^\circ \times 0.4^\circ$, domain.

The planetary boundary layer (PBL) was parameterized with the Eta Mellor–Yamada 1.5-order local closure scheme (Janjic 1990, 1994), which includes a prognostic equation for turbulence kinetic energy (TKE). Betts–Miller cumulus parameterization (Betts 1986; Betts and Miller 1986) was used for D1 and D2 only, while cloud-microphysical processes were explicitly resolved in all domains using the Reisner mixed-phase scheme (Reisner et al. 1998). Shortwave and longwave radiation were computed using the rapid radiation transfer model (RRTM; Mlawer et al. 1997).

A second simulation was run to assess the frictionally forced ascent without the effects of vertical shear using the model described by Kepert and Wang (2001), with some minor modifications to suit this case. This dry,
hydrostatic model of the tropical cyclone boundary layer diagnoses the boundary layer flow as the response to certain forcing boundary conditions. The upper boundary forcing is a translating pressure field representative of a tropical cyclone. The parametric pressure field of Willoughby et al. (2006) was used for this purpose, with the adjustable parameters chosen by a minimized least squares fit to the MM5 mean sea level pressure field (Fig. 3). The development of this parametric model included very extensive verification against aircraft data, and it has also been successfully used in analyses of the tropical cyclone boundary layer by Kepert (2006a,b) and Schwendike and Kepert (2008). The surface boundary condition consists of a surface roughness calculated by a Charnock relation over the sea, and a fixed roughness length over land. For these simulations, the Charnock coefficient was increased from 0.011 to 0.018 to match the MM5 parameter, and the land roughness length of 0.5 m similarly matches MM5. The land is assumed to be flat, with the coastline approximated by a straight line oriented north–south, and the land surface and sea surface temperatures are set at 300 K. For simulations not involving landfall, the model integrates the equations of motion with these forcings until a quasi–steady state is attained. As TC Larry made landfall during the period of interest, equilibrium is not attained, but the landfall time was set to occur at 24 h, which is sufficient to ensure that the modeled boundary layer has fully spun up. The model includes a choice of parameterizations for the surface layer and turbulent closure. For these calculations, the surface layer parameterization used Monin–Obukhov similarity theory with the momentum roughness given by a Charnock (1955) relation and the heat and moisture roughness lengths after Liu et al. (1979), with a Mellor–Yamada level-2.25 closure (Galperin et al. 1988) for levels above that. These parameterizations are similar to those used for the MM5 runs, and it is not expected that the small differences will have a material effect on the results.

Measuring the cyclone axis tilt requires that the center be accurately found. The center may be defined in several different ways, in terms of either the wind or the mass fields, and the most appropriate method may be application dependent. Kepert (2005) reviews and compares some of the current techniques. While most of the analysis of this problem has been for observations, the techniques are also applicable to model output. We prefer to define the center in terms of the mass field, since boundary layer inflow can cause problems with wind-based methods near the surface (Kepert 2005). The obvious choice is the point of lowest pressure, for which we first apply low-pass spatial filtering with a cutoff of 5 km below 500 hPa and 8 km above to remove small-scale noise. This technique tends to give the location of the strongest small-scale vortex within the eye, if such are present, rather than the geometric center of the maximum wind belt, and so we refer to it as the “eye scale” track. Alternatively, one could fit a parametric tropical cyclone pressure profile, adjusting the profile structure and position to optimize the fit. The implementation of this idea was described in the context of observational analysis by Kepert (2005), who showed that the technique was sensitive to larger scales within the cyclone than simply choosing the lowest pressure, since the parametric profile filtered out features such as eyewall mesovortices and vortex Rossby waves. Specifically, we do an iterative nonlinear least squares fit of a Holland (1980) profile to the model mass data, adjusting the vortex intensity, RMW, “B parameter,” and location. Data are used out to a radius of 50 km from the point of lowest pressure, with the results being fairly insensitive to other choices in the range of 15–80 km. Since this method is sensitive to larger scales, we refer to it as the “vortex scale” track. The axis tilt follows from applying the method at several different heights in the model.

3. Results
The MM5 TC track and intensity are in good agreement with the observed track and intensity of TC Larry,
including a period of rapid intensification, as shown in Fig. 4. After insertion of the initial bogus vortex and allowing sufficient time for vortex spinup (~24 h), the simulated TC track matches closely with the observed track from about $t = 30$ h to decay over land at $t = 72$ h (Fig. 4a). The overall trend in the simulated central pressure also exhibits a similar life cycle to the observed TC, with steady intensification from $t = 0$ h to $t = 24$ h, rapid intensification from $t = 36$ h to $t = 57$ h, followed by a pronounced decay after landfall at $t = 59$ h (Fig. 4b).

a. Rainfall characteristics

The accumulated rainfall pattern in the 12 h from +48 to +60 h (11 h before landfall to 1 h after landfall) is highly asymmetrical with maximum values, occasionally exceeding 300 mm, located on the right side of the TC relative to its westward direction of motion (Fig. 5a).
Much lower rainfall totals occur on the left side of the track, with the exception of isolated areas of greater than 200 mm associated with spiral bands. These spiral bands appear to form quasi-periodically on the northern side of the storm, or to the right of motion. Animation of cloud rainwater mixing ratio in D3 and D4 shows that these bands rotate cyclonically around the rear side of the vortex, as evidenced by the streaky pattern of accumulated rainfall to the left of the TC track in Fig. 5. In the 4 h from +54 to +58 h almost twice as much rain falls to the right of track than to the left of track (Fig. 5b). Over this same time window, the central

Figure 5. (a) The 12-h (+48 to +60 h) accumulated explicit precipitation from D3 (gray shading with contour interval of 50 mm). Solid black circles indicate the position of the TC center every hour from +48 to +60 h with the period from +54 to +58 h marked by the black line. The 200- and 300-mm contours are outlined in white. The dashed line marks the eastern boundary of D4. The abscissa and ordinate are grid cell number. (b) As in (a), but for the 4-h accumulated precipitation from +54 to +58 h with a contour interval of 25 mm.
pressure decreases from 933 hPa at +54 h to a minimum central pressure of 929 hPa at +57 h, followed by a slight increase to 930 hPa at +58 h as the TC nears landfall (Fig. 4b). Although this window of time is only a third of the 12-h period of accumulated rainfall shown in Fig. 5a, we found that the asymmetries in low-level vertical motion and convergence (discussed in detail in section 3b) that result in such a pronounced rainfall asymmetry are qualitatively persistent over the 12 h.

Figure 6 shows instantaneous values of rainwater mixing ratio $q_r$ at the surface and 5000 m, and vertical velocity at 2000 m, each hour from +54 to +57 h and every 12 min from +55 to +56 h. Analyses of the three-dimensional structure of $q_r$, suggests that precipitation originating from convection in the TC core falls in a helical manner, owing to strong cycloidal advection of hydrometeors. While the largest $q_r$ values at the surface are located persistently on the right side of the core circulation, in agreement with the accumulated rainfall pattern shown in Fig. 5b, the maximum $q_r$ values at 5000 m occur upstream from the surface maximum, with angles ranging from 30° to 120° (Fig. 6). A three-dimensional image of one of these helical precipitation cores, as resolved by D4, is presented in Fig. 7. The midlevel maxima in $q_r$ are generated by deep convection that will be seen to be forced by strong convergence in the boundary layer ahead of the TC. The vertical velocities at 2 km, associated with this region of persistent convergence, are generally between 2 and 5 m s$^{-1}$. At times, the vertical vorticity within the eyewall forms localized maxima, which are more discrete, and have associated low-level vertical velocities as high as 8 m s$^{-1}$. During such times, the low-level reflectivity structure around the eye assumes polygonal-like shapes such as squares and pentagons (not shown here), in agreement with several previous studies (e.g., Kossin and Schubert 2001; Kossin et al. 2002; Braun et al. 2006). These transient features tend to increase significantly the value of $q_r$ in the midlevels of the eyewall region resulting in locally very heavy precipitation. Evidence of such transient bursts in eyewall convection can be seen by the quasi-periodic localized maxima of accumulated rainfall to the right of track shown in Fig. 5b.

b. Low-level vertical motion, vorticity, and convergence

The strong and persistent asymmetry in the accumulated rainfall pattern suggests a persistent underlying asymmetric forcing mechanism associated with the inner-core convection. Indeed, an examination of low-level vertical motion, relative vorticity, and surface convergence over the 5 h from +54 to +59 h, using data from D3 and D4, reveals persistent asymmetrical structures that help to explain the rainfall distributions shown in Fig. 5.

The mean vertical motion at 1-km height from the MM5 simulation is shown in Fig. 8. The mean updraft (Fig. 8a) consists of two main features: a strong semicircle located slightly to the left of front on the RMW, with an attached spiral band extending from this arc to the south, and a similar band extending northward from the rear of the storm. Examination of the standard deviation of w reveals that the eyewall arc and southern band have much less temporal variability than the northern band (Fig. 8d). Azimuthal Fourier decomposition of $w$ reveals that the wavenumber-1 component has peak amplitude of 2.3 m s$^{-1}$ close to the radius of maximum symmetric updraft and within the RMW (Fig. 8b). In combination with the symmetric component, which peaks within the eyewall at 1.7 m s$^{-1}$, this accounts for both the strong updraft at the left front of the eyewall and the weak descent at the rear (Fig. 8e). The wavenumber-2 component has about half the maximum amplitude of wavenumber 1, but this peak occurs outside of, rather than at, the RMW (Fig. 8c). This component accounts for a portion of the spiral bands extending to the north and south (Fig. 8f).

A simulation of the frictionally forced updraft, as diagnosed by the boundary layer model of Kepeht and Wang (2001), is shown in Fig. 9. This simulation was forced by a parametric profile after Willoughby et al. (2006), which was fit to the MM5 surface pressure field averaged over the period +56 to +59 h with the storm motion removed using the surface vortex-scale track. The fitted profile is shown in Fig. 3. Six hours before landfall, the main updraft occurs in a semicircle centered on the front left of the eyewall, with a peak of about 2.5 m s$^{-1}$, and descent of −1.5 m s$^{-1}$ to the right rear. A band of ascent to the north begins to appear shortly afterward, and is prominent by 2 h afterward, or 4 h before landfall in the simulation (Fig. 9b). A weaker band extending to the south from the eyewall maximum also forms at about this time. These bands strengthen, most obviously the northern one, which also migrates toward the front of the storm as it approaches land (Figs. 9c,d). Fourier decomposition of the vertical motion field revealed similar contributions from the symmetric and lower-wavenumber components to those of the MM5 simulation, with good agreement in both phase and magnitude (not shown). A simulation without landfall did not produce the northern and southern bands, with the vertical motion field being similar to that in Fig. 9a throughout (not shown). Since (i) only minor changes to the updraft at the left front of the eyewall are seen with the approach to land, (ii) the location and structure of this feature are similar to the
simulation without landfall and to the results of earlier nonlandfall studies, and (iii) this feature is dominated by azimuthal wavenumber 1, we conclude that this feature is the result of asymmetric friction due to the storm motion, similar to that discussed in Shapiro (1983), Frank and Ritchie (1999), Kepert (2001), Kepert and Wang (2001), and Kepert (2006a). The northern band is attributed to the effects of the land–sea roughness contrast,
which apparently sets up a flow perturbation that propagates or is advected toward the storm center, similar to that found in the observational and modeling analysis of Hurricane Mitch (1998) when it was a similar distance from the Honduras coast by Kepert (2006b). The cause of the weaker, southern band is less clear, but is presumably some kind of dynamical adjustment of the circulation to the intense northern band.

The vertical motion near the boundary layer top in this dry diagnostic model displays strong quantitative and qualitative similarities to that in the MM5 simulation. The principal differences are that the MM5 simulation
shows less tendency for the northern band to rotate with time, has more small-scale structure due to the effects of moist convection, and has some topographic forcing of ascent over the land. The boundary layer model does not contain any moist physics, nor does it contain any asymmetric forcing of ascent due to environmental shear. We are therefore confident that the dominant forcing of the vertical motion in the lower part of the MM5 simulation is from the surface, being a combination of symmetric frictional forcing, a strong wavenumber-1 asymmetry maximized at the eyewall due to motion-induced asymmetric friction, and a higher-wavenumber asymmetry outside the eyewall due to the land–sea frictional contrast that grows as the storm approaches land.

Strong low-level relative vorticity values at 1-km height, with values approaching $-1.6 \times 10^{-2}$ s$^{-1}$, are found predominantly in the front and front-right quadrants of the TC eyewall in the MM5 simulation (Fig. 10). These concentrated regions of eyewall vorticity tend to form in association with locally strong gradients of azimuthal wind speed near the RMW, and extend vertically to about 2 km. At times, these low-level relative vorticity maxima tend to congeal into discrete pools such as at $t = +54$ h (Fig. 10a), whereas at other times the vorticity becomes elongated in a narrow bands along the front quadrants of the eyewall where surface convergence is also maximized (Figs. 10b,c). Cross-sectional analyses though these features suggest that they sometimes are connected with very strong vertical velocities within the eyewall of up to 18 m s$^{-1}$ in the mid- to upper troposphere (not shown). However, because stretching is one of the major contributing terms to changes in vertical vorticity it is unclear as to whether these low-level vorticity maxima are forming in response to strong updrafts ($>10$ m s$^{-1}$) within the eyewall or are forcing the updrafts. Furthermore, the maximum boundary layer convergence and deep convective structures are predominantly located on the trailing edge of the vorticity pools, suggesting that, as Braun et al. (2006) argued, the most intense convective towers result from the interaction between mesovortices in the eyewall and low-level storm-relative inflow.

Another distinct and persistent asymmetry is noted in the surface convergence field, with maximum values most often located in the front quadrants of the TC (Figs. 10e,h)—consistent with the mean vertical motion at 1 km shown in Fig. 8a. An analysis of the 4-min output data from D3 and D4 indicates strong surface convergence forms in a broad band ahead of the TC, sometimes intensifying in the front-right quadrant of the eyewall in association with localized low-level vorticity maxima. The maximum surface $q_r$ is displaced typically by about 90° to the right of the direction of the maximum surface convergence because of the rapid advection of hydrometeors by the tangential winds in the TC core (Fig. 7). This angle appears to be somewhat
sensitive to the updraft intensity, with a greater (lesser) separation between precipitation source and sink regions for weaker (stronger) updrafts.

c. Storm motion, vertical shear, and vortex tilt

The translation speed of the simulated TC is in good agreement with the observed translation speed of TC Larry in the 7 h leading up to landfall, with their respective tracks almost overlapping (Fig. 4a). Both the simulated and observed TC moved westward at a speed of about 7–7.5 m s\(^{-1}\). The deep-layer-averaged vertical shear of the environmental wind is particularly low during the 5-h period from \(t = 53\) to \(t = 58\) h (Fig. 11). At \(t = 54\) h the shear is from the south-southeast (163°) with a magnitude of 2.7 m s\(^{-1}\), before subsequently veering to a more southerly direction (−180°) and gradually weakening (Fig. 11). By \(t = 57\) h the shear decreases further to about 2.5 m s\(^{-1}\) during which time the TC is near its minimum central pressure of 928 hPa. Despite the overall weak environmental shear (<3 m s\(^{-1}\)), the simulated TC maintains its asymmetric accumulated rainfall pattern until landfall when orographic and land surface friction effects come into play. The local shear, calculated over the same time period, is slightly stronger, ranging from 3 to 5 m s\(^{-1}\), and interestingly is opposite in direction to the environmental shear—that is, from the north-northwest (not shown). These results suggest that for intense TCs embedded in relatively weak shear, the storm translational speed can be the dominant factor in determining the spatial distribution of convective rainfall in the TC core, in agreement with some previous observational (e.g., Corbosiero and Molinari 2003; Lonfat 2004; Chen et al. 2006) and modeling (Frank and Ritchie 1999) studies.

The vortex-scale track of the surface center of the simulated TC in D4, together with the mean tilt from this method, are shown in Figs. 12a,b. There is a clear and vertically coherent axis tilt to the south then east, but of less than 1 km, from this method. This tilt is thus consistent in direction with the local shear, but is opposite to the environmental shear. The small magnitude of the tilt is qualitatively consistent with the magnitude of the shear. Furthermore, there is no evidence of precession of this tilt, contrary to some other simulations of TCs in vertical shear (e.g., Braun et al. 2006). Figure 12 also shows the eye-scale tracks at 850 and 200 hPa. These tracks display a clear cycloidal character, and are well approximated by a cyclonic circular oscillation.
superimposed on the much smoother vortex-scale track (Fig. 12c). At 850 hPa this oscillation has a radius of 1–2 km and a period of about 30 min, increasing to about 5 km and 45–50 min at 200 hPa. The angular period of the eyewall winds is about 30–35 min near the surface, increasing to 60–70 min at 200 hPa, so while the eye-scale center is rotating with the winds at low levels, it apparently lags them slightly in the upper part of the storm, possibly because it is rotating with the winds at a slightly lower level. Examination and Fourier decomposition of the individual pressure fields showed that the cycloidal eye-scale track at these levels was due to the
rotation of a wavenumber-1 feature contained within the eye. Interestingly, at 500 hPa the dominant asymmetry had azimuthal wavenumber 2 rather than 1, and the eye-scale center finder chose one of these essentially at random, so a cycloidal track was not apparent at this level. The pressure minima associated with the cycloidal eye-scale track were closely located with similarly rotating cyclonic vorticity anomalies. The small-amplitude trochoidal oscillation shown here is very similar to the wavenumber-1 instability and associated "wobble" of the TC eye first elucidated by Nolan and Montgomery (2000), and further explored by Nolan et al. (2001) using a highly idealized, dry simulation of a hurricane-like vortex. In particular, our low-level vorticity anomaly here is confined within the eye, rotates with the mean angular velocity, and is only a few kilometers deep. However, the cause of the oscillation at 200 hPa remains unknown.

The symmetric vorticity structure of the storm is shown in Fig. 13. All levels show essentially the same structure: a ring of elevated vorticity just inside the RMW, with lower values at the storm center. The strength of the ring decreases with height, consistent with the decrease in wind speed and expansion of the eye as height increases. This structure supports the instability analyzed by Nolan and Montgomery (2000) and Nolan et al. (2001), which presumably accounts for the formation of the low-level wavenumber-1 vortex that results in the cycloidal eye-scale track discussed above. It also satisfies the Schubert et al. (1999) necessary condition for instability, consistent with the vorticity pooling shown in Fig. 10a. Outside this hollow tower, the vortex displays a "skirt" of vorticity similar to that seen in observations (Mallen et al. 2005). Such a structure has been argued by Reasor et al. (2004) and Schecter et al. (2002) to make the vortex resilient to tilt (i.e., any tilt of the vortex axis due to, say, environmental shear, will decay with time). This structure is therefore consistent with the very small tilt found using the vortex-scale track at various levels, and indirectly supports our finding that the vertical motion asymmetry in the lower part of the storm is dominated by asymmetric frictional forcing. Based on the results presented in this section, we can conclude that significant inner-core asymmetries are possible even when the TC vortex has little to no mean tilt.

4. Discussion and conclusions

Severe TC Larry (2006) was simulated with the MM5 model, using high-resolution nested domains with grid spacings of 3 (D3) and 1 km (D4) to resolve detailed inner-core structures. Additional simulations were carried out using the boundary layer model described by Kepert and Wang (2001) to assess the frictionally forced ascent without the effects of vertical shear. Comparisons between the MM5-simulated TC, best-track data, and satellite/radar imagery, indicated that the model was able to reproduce the main characteristics of TC Larry, including its track, translation speed, intensity, and size. The central pressure, as resolved by the 3-km grid, was only 1–2 hPa above that obtained with the 1-km grid, suggesting that the TC eye for this storm is well resolved for grid spacings below about 3 km. The inner-core structures in D4 were qualitatively similar to those in D3 in terms of location with respect to the eye; however, individual features such as low-level vorticity and vertical velocity were greater in magnitude, as expected. Furthermore, because D3 covered a much greater geographical area than D4 it allowed us to assess the temporal continuity of the TC asymmetries for a period of at least 12 h to match the accumulated rainfall pattern presented in Fig. 5a.

A distinct asymmetry was found in the accumulated rainfall pattern during the 12 h prior to landfall, with the heaviest precipitation occurring along a swath to the right of the storm track and about 15 km from the TC center. This asymmetry matched closely with the both the coldest cloud-top temperatures from microwave satellite imagery (Fig. 1) and regions of heavy precipitation as detected by Bureau of Meteorology’s radar network (not shown). The rainfall pattern is consistent with the rainfall climatology for TCs suggested by Lonfat et al. (2004), who showed that the asymmetry in the rainfall maximum moves from the front-right quadrant for categories 1–2 TCs to the front-left quadrant for categories 3–5 TCs (in the Northern Hemisphere). For the SH this implies a corresponding shift.
from the front-left quadrant to the front-right quadrant for severe TCs such as TC Larry. Our results are in agreement with the climatological study of Lonfat (2004) and Chen et al. (2006), who showed that for TCs in the SH the inner-core rainfall asymmetry tends to be downshear and to the right, although this would only be true for the environmental shear here. Lonfat (2004) and Chen et al. (2006) showed also that this rainfall asymmetry is about twice as strong for fast-moving TCs (>5 m s\(^{-1}\)) than for slow-moving TCs (<5 m s\(^{-1}\)). While we have not explored the sensitivity of rainfall asymmetry to translation speed (because we are only considering one case), both the observed speed of TC Larry and the simulated speed of about 7.5 m s\(^{-1}\) are relatively high when compared with the climatological average speed of TCs in the Coral Sea region of between 2.5 and 5 m s\(^{-1}\) (Chen et al. 2006). Furthermore, the relatively low vertical wind shears of less than 3 m s\(^{-1}\) over both the shear environment and the TC core region, in addition to almost no mean tilt of the modeled vortex during the time of analysis, strongly suggests that translation-induced frictionally forced ascent is the dominant factor in determining the asymmetries present.

This argument is further supported by the simulations carried out using the Kepert–Wang boundary layer model with the surface parameters adjusted to match the MM5 simulation. A Fourier decomposition of the low-level vertical motion produced by this model reveals a dominant, positive, wavenumber-1 component at the left front of the eyewall. This feature is consistent with the low-level vertical motion structure produced by the MM5, with good agreement in both phase and magnitude. The semistationary band of ascent on the northern side of the MM5-simulated TC is dominated by a wavenumber-2 asymmetry, and is only reproduced by the boundary layer model when the effects of land are included. The boundary layer model is able to reproduce the main features of the low-level vertical motion in MM5 with no forcing due to environmental shear. Therefore, we conclude that the dominant asymmetric forcing in the MM5 simulation is from the surface.
The pronounced rainfall asymmetry can be explained physically by enhanced surface convergence and low-level vertical motion ahead of the translating TC, which results in deep convection and precipitation in the midlevels. The maximum surface rainwater tends to be displaced by about 30°–120° to the right of this low-level convergence maximum, owing to strong cyclonic advection of hydrometeors by the winds in the TC core. Shapiro (1983), Kepert (2001), and Kepert and Wang (2001) have shown that TC-like vortices developed asymmetries in the boundary layer forced by the asymmetry in frictional drag caused by the translation of the storm, such that maximum convergence and vertical motion occurs to the front or right-front of the storm (in the Northern Hemisphere). Our results suggest this is the primary mechanism governing the rainfall asymmetry for this case. The direction of the environmental vertical shear vector relative to the rainfall maximum is inconsistent with some recent case studies for Northern Hemisphere TCs. Braun and Wu (2007) showed the maximum upward motion and rainfall tend to occur to the left of the shear vector—that would be to the right of the shear vector for a Southern Hemisphere storm. We show that the maximum upward motion in our simulation also occurs to the left of the environmental shear vector. However, the relatively low shear magnitude combined with the very small mean vortex tilt (~1 km) suggests that the environmental shear has only a minor effect on the asymmetries of TC Larry when compared to the effects of the translation speed. Furthermore, the direction of vortex tilt is opposite to the direction of the environmental vertical shear vector but is in the same direction as the local shear, suggesting that the tilt, while small, is possibly a response to shear confined only to the inner-core region of the vortex. If so, this implies a degree of complexity in the interpretation of the effects of shear that warrants further investigation.

We have shown that significant inner-core asymmetries are possible in environments with relatively weak deep-layer vertical wind shear, complementing previous studies in which large environmental vertical shear has been shown to play a dominant role in determining TC asymmetries (e.g., Rogers et al. 2003; Braun and Wu 2007). When the environmental vertical shear is low, as in the current simulation, the asymmetric boundary layer convergence caused by the translation of the TC itself can become the primary mechanism forcing the inner-core asymmetries, in accordance with previous theoretical and modeling studies.

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